Mesoscale Spectra of Mars’s Atmosphere Derived from MGS TES Infrared Radiances

TAKESHI IMAMURA
Institute of Space and Astronautical Science, Japan Aerospace Exploration Agency, Sagamihara, Kanagawa, Japan

YASUHIRO KAWASAKI
IP Development and Management Division, Hitachi, Ltd., Tokyo, Japan

TETSUYA FUKUHARA
Institute of Space and Astronautical Science, Japan Aerospace Exploration Agency, Sagamihara, Kanagawa, Japan

(Manuscript received 10 April 2006, in final form 1 September 2006)

ABSTRACT
Wavenumber spectra of the atmospheric potential energy of Mars at mesoscales (wavelengths of 64–957 km) were obtained as a function of latitude, season, and Martian year using infrared radiance data obtained by the Thermal Emission Spectrometer (TES) onboard the Mars Global Surveyor (MGS) spacecraft. Spectral slopes tend to be flatter at smaller scales, and the slopes are usually flatter than $-1$ near small-scale ends. Near large-scale ends, the spectra sometimes show prominent steepening with slopes from $-2$ to $-3$. The power peaks in the high latitudes in winter and equinoxes, suggesting that eddies are generated preferentially in baroclinic zones. The seasonal variation at each latitude band, on the other hand, tends to be obscured by large interannual variability. An enhancement in the power was observed around the storm tracks in the Southern Hemisphere. Spectra of the terrestrial stratosphere were also obtained with a similar method from data taken by the Aura satellite and compared to the results for Mars.

1. Introduction
Atmospheric energy spectra as a function of horizontal wavenumber give clues as to the energy cycles of planetary atmospheres. The standard view for the terrestrial troposphere begins with the generation of zonal available potential energy by differential solar heating. This is converted to eddy available potential energy and eddy kinetic energy via baroclinic instability, principally in zonal wavenumbers 2–10 (e.g., Koshyk and Hamilton 2001). Nonlinear interactions transfer the eddy energy both upscale and downscale from the generation scale. Energy injection at smaller scales via moist convection and the generation of internal gravity waves will also contribute to the spectra.

Nastrom and Gage (1985) analyzed wind and temperature data of the terrestrial atmosphere taken from commercial airplane flights and obtained kinetic and potential energy spectra as a function of horizontal wavenumber. Each spectrum has two different wavenumber regions with distinct power-law dependencies. At longer wavelengths of $>600$ km the spectra agree with previous analyses using global datasets, which showed approximately $k^{-3}$ power law where $k$ is the horizontal wavenumber (e.g., Boer and Shepherd 1983). This power law is in accord with the nature of enstrophy-cascading inertial range in two-dimensional turbulence. The resemblance between the observed spectra and the spectra for two-dimensional turbulence is attributed to the quasigeostrophic nature of synoptic-scale motions (Charney 1971). Although quasigeostrophic flow is not strictly two-dimensional, Charney (1971) argued that the $-3$ slope is appropriate for quasigeostrophic turbulence and that energy should be partitioned equally among each of the two horizontal velocity components and potential energy. The energy source of the turbulence will be forcing arising from baroclinic instability, as the energy increases in the mid-
and high-latitude baroclinic zones and increases in winter (Nastrom and Gage 1985).

The spectra by Nastrom and Gage (1985) additionally show a $k^{-5/3}$ power-law regime at mesoscales with wavelengths of 2–600 km. The interpretation of this part of the spectra has been controversial (e.g., Koshiy and Hamilton 2001). Lilly (1989) argued that energy injected by small-scale convective activity cascades upward in stratified two-dimensional turbulence and creates a $-5/3$ slope. On the other hand, Cho and Lindborg (2001) showed the occurrence of downward energy cascade based on the structure function analysis of wind and temperature data taken from aircraft. Recent numerical models reproduce downward energy cascade and a $-5/3$ slope at mesoscales (Tung and Orlando 2003; Skamarock 2004; Lindborg 2005). On the other hand, VanZandt (1982) suggested that motions in this regime contain a significant contribution from internal gravity waves. The enhancement of mesoscale variances over mountainous terrain was observed from airplane flights and attributed to topographically generated waves (e.g., Jasperson et al. 1990). Studies with numerical models support these suggestions; divergent components have spectral slopes much flatter than $-3$ and contribute with magnitudes comparable to rotational components at small scales, accounting for the shallowing of the spectra at mesoscales (e.g., Yuan and Hamilton 1994; Koshiy and Hamilton 2001; Kitamura and Matsuda 2006). All these processes might work together to fill the mesoscale energy spectrum (Skamarock 2004).

This study is the first to investigate the wavenumber spectra of the Martian atmosphere at mesoscales, aiming at characterizing the statistical nature of disturbances and getting insight into the atmospheric energy cycle. Mars has an equatorial radius of 3394 km (i.e., about half the size of Earth) and rotates with 1 solar day (sol) of 24.66 h and an inclination of the axis of 25.19°. The mean distance to the sun is 1.52 times the Sun–Earth distance with an orbital period of 686.98 Earth days or 668.60 sols. The atmospheric composition is mainly CO$_2$ with an average surface temperature of 210 K (e.g., Kieffer et al. 1992). A large amount of dust floats in the atmosphere, giving the globally averaged dust opacities of 0.05–0.4, and H$_2$O ice clouds also provide highly variable cloud opacities (e.g., Smith 2004). Carbon dioxide ice clouds are thought to exist in the polar night (e.g., Pearl et al. 2001). The height of the convective boundary layer is considered to be 4–5 km based on temperature measurements and numerical modeling (e.g., Toigo et al. 2003).

The latitudinal temperature gradient of the Martian atmosphere is extremely large in the high latitudes of the winter hemisphere because of the strong radiative cooling near the Pole, while temperature is horizontally close to uniform in the summer hemisphere (e.g., Banfield et al. 2003). In the regions of strong baroclinicity, baroclinic instability will generate disturbances with zonal wavenumbers 2–4 (e.g., Barnes 1980, 1981, 1984), which cascade into smaller-scale turbulent eddies via nonlinear processes. Given the buoyancy frequency of $N = 0.006–0.01$ s$^{-1}$, Coriolis parameter at 45° latitude of $f = 1 \times 10^{-4}$ s$^{-1}$, and scale height of $H = 10$ km, the deformation radius is $NH/f = 600–1000$ km, which is similar to Earth’s; the most unstable wavelength will also be similar, with a smaller zonal wavenumber due to the smaller planetary radius. Westerly winds in baroclinic zones flowing over topography will generate internal gravity waves as well (Barnes 1990). The nature of the mesoscale eddies on Mars is an important issue also from the viewpoint of diffusive dust transport, which may enhance dust storms.

The data used are the radiances of upwelling thermal radiation in the CO$_2$ 15-μm band measured by the Thermal Emission Spectrometer (TES) onboard the Mars Global Surveyor (MGS), hereafter referred to as the MGS TES. MGS was launched in November 1996 and inserted into a Martian sun-synchronous polar orbit in March 1999, which is the start of the mapping phase. MGS passes over the equator at approximately 0200 and 1400 local time (LT) and provides spatially dense sampling in a roughly north–south direction along orbits; these one-dimensional radiance data are converted to wavenumber spectra of atmospheric energy. Section 2 describes the characteristics of the dataset, section 3 describes the method of calculating spectra, and section 4 provides results. In section 5, wavenumber spectra of the terrestrial stratosphere are obtained with a similar method and compared with the spectra of the Martian atmosphere. Section 5 summarizes the conclusions.

2. Dataset

The MGS TES is a Fourier transform spectrometer that measures thermal radiation at wavenumbers 200–1700 cm$^{-1}$ (Christensen et al. 2001). The data obtained by the MGS TES are distributed through the National Aeronautics and Space Administration (NASA) Planetary Data System (PDS). We analyze the MGS TES brightness temperatures at the center of the CO$_2$ 15-μm band (668.74 cm$^{-1}$) measured in nadir geometry with emission angles $<2°$. The contribution function at this wavenumber peaks at $\sim$23-km altitude ($\sim$0.5 hPa) with the full width at half maximum (FWHM) of $\sim$18 km, while
the contribution from altitudes below 10 km (~2 hPa) is negligible (see Conrath et al. 2000). With this contribution function, temperature disturbances with vertical wavelengths of 20, 40, and 80 km will produce brightness temperature fluctuations with amplitudes 0.08, 0.40, and 0.74 times the original amplitudes, respectively. The altitudes of Olympus Mons, Tharsis Montes, and Elysium Mons exceed 10 km; since the emission from such highlands might influence the 668.74 cm
brightness, data taken near these regions (211°–260°E 
× 15°–25°N and 140°–150°E × 20°–30°N) are excluded from the analysis. Within the emission angle of 
<2°, brightness temperatures observed should not differ from those from exact nadir geometry by more than 0.01 K. Detector-2 data are used from among the data taken by 6 identical detectors mounted on TES. Only dayside (1400 LT) data are used in this study; comparison with the night side is left for future studies.

It would be possible to use atmospheric temperatures that are retrieved from radiance spectra and provided through PDS instead of brightness temperatures. However, retrieved temperatures sometimes contain errors caused by inappropriate assumptions in the retrieval algorithm, and such errors might influence the present analysis that focuses on small-amplitude fluctuations. We found that retrieved temperatures at 0.5 hPa sometimes give much larger power than brightness temperatures at small scales near the equator, although retrieved temperatures and brightness temperatures usually yield similar results. The influence of the measurement error in radiances on wavenumber spectra is discussed in section 4.

The MGS mapping phase gives observations with latitudinal resolutions of up to ~0.1° (~6 km) along 12 orbits within 1 sol, and these orbits are spaced roughly 30° apart in longitude. The dense sampling in latitude enables us to calculate mesoscale spectra as a function of meridional wavenumber. The wavenumber spectrum does not depend on the direction of sampling when atmospheric motion is horizontally isotropic; this is true of the motions at the total spherical harmonic index of >8 in the terrestrial troposphere (Shepherd 1987). The present study focuses on meridional wavenumbers 22.5–360, which are expected to be within the isotropic regime. Internal gravity waves, however, will have anisotropy depending on the direction of background wind.

The data used cover a period from February 1999 to August 2004, which is within the mapping phase. In terms of the areocentric longitude of the Sun ($L_S$) and Mars year (MY) (Clancy et al. 2000; Smith 2004), this period spans from $L_S = 104°$ in MY 24 to $L_S = 81°$ in MY 27. The $L_S$ is a measure of seasons: $L_S = 0°$ at the spring equinox in the Northern Hemisphere, $L_S = 90°$ at the summer solstice, $L_S = 180°$ at the fall equinox, and $L_S = 270°$ at the winter solstice.

3. Analysis procedure

Wavenumber spectra in five latitudinal bands were obtained as a function of season and Mars year through the procedure described below.

a. Preparation of one-dimensional data

The MGS TES has two different modes for the scan length in taking interferograms: single-scan mode (FWHM of spectrometer line shape ~13 cm
and double-scan mode (FWHM ~6.5 cm
). These modes are sometimes switched with each other during the mission period. To avoid the possible influence of mode changes on the result, the double-scan radiances at wavelengths of 663.42 cm
channel 98), 668.74 cm
channel 99), and 674.06 cm
(channel 100) are averaged with weights of 0.25, 0.5, and 0.25, respectively, and the resultant radiances are considered equivalent to the single-scan radiance at 668.74 cm
(channel 50). Furthermore, orbits in which mode changes occur are excluded from the analysis.

Observed radiances sometimes include spiky offsets with magnitudes comparable to the typical radiance values. These offsets are regarded as artificial, and the data showing such features are excluded. The radiances during MY 24, $L_S = 122°–125°$ sometimes drop abruptly by more than 10 K in brightness temperature. Such structures are also unlikely to be real, and thus the data during this period are not used. Formally, the noise level of each observation is given by the “spectrometer noise” index in the database. If the data whose spectrometer noise indexes are not “nominal level” are excluded, the number of data used is much reduced during MY 26, $L_S = 90°–180°$ and MY 27, $L_S = 60°–90°$, and consequently, the spectra averaged for these periods become rather noisy. In spite of such an increase in noise, however, spectral densities and slopes are basically not changed. Therefore, noises designated by this index are not considered serious, and the index is not used for data selection.

The brightness temperatures of observed radiances are regarded as atmospheric temperatures near 23 km (section 2). These temperatures along orbits are binned in latitude increments of 0.5°, with averaging in each grid (Fig. 1). The grids at which observations do not exist are filled with linear interpolations of the nearest observation points only when the interval to be interpolated does not exceed 2°. Changes of this threshold to 1° and 4° do not change the basic features of the result, suggesting a minor effect of the interpolation.
The binned data along each orbit are divided into 5 latitudinal segments of 32° long centered at 64° S, 32° S, 0°, 32° N, and 64° N. Each segment usually includes 64 data points; segments with less than 64 are not used. Hereafter these 5 latitudinal bands are denoted simply by their central latitudes. The data are further classified into 12 seasons with periods of 30° in LS.

b. Calculation of wavenumber spectra

Before calculating spectra, the mean and a linear trend are removed from each latitudinal segment (Fig. 1). The residual is converted to a power spectrum by digital Fourier transform with Welch window function that drops to zero near both ends of the latitudinal segment. Subtraction of a second-order function instead of a linear trend does not change the spectral densities significantly except in the gravest Fourier component, which is excluded from the result.

The high-latitude segments centered at 64° S and 64° N sometimes include distinct latitudinal structures associated with polar jets. These structures tend to remain after subtracting linear trends and lead to misleading results. To suppress this effect, each 32° high-latitude segment is divided into two 16° segments, which are spectrally analyzed separately with the procedure above. The resultant two spectra are averaged to get a spectrum for 64° S or 64° N, with the maximum wavelength in the spectrum being reduced by half.

The spectra obtained are averaged in each latitudinal band and season. Figure 2 shows the numbers of the spectra used for averaging as a function of season, for both single-scan and double-scan observations. Note that the numbers for 64° S and 64° N were almost doubled by the division into two segments.

The averaged temperature spectra are converted to potential energy spectra with a method described by Gage and Nastrom (1986). A temperature spectrum \( \Phi_{TT} \) is related to the potential temperature spectrum \( \Phi_{\theta\theta} \) as

\[
\Phi_{TT} = \frac{T^2}{\bar{T}^2} \Phi_{\theta\theta},
\]

where \( T \) is the background temperature and \( \bar{T} \) the background potential temperature. Assuming that the horizontal gradient of \( \bar{T} \) is small, \( \Phi_{\theta\theta} \) is related to the vertical displacement spectrum \( \Phi_{\zeta\zeta} \) through the vertical potential temperature gradient \( d\bar{T}/dz \) or the buoyancy frequency \( N \):

\[
\Phi_{\theta\theta} = \left( \frac{d\bar{T}}{dz} \right)^2 \Phi_{\zeta\zeta} = \frac{\bar{T}^2}{g^2} N^4 \Phi_{\zeta\zeta},
\]

where \( g = 3.72 \text{ m s}^{-1} \) is the gravitational acceleration. The \( \Phi_{\zeta\zeta} \) is further related to the potential energy spectrum \( \Phi_{PE} \) as

\[
\Phi_{PE} = \frac{1}{2} N^2 \Phi_{\zeta\zeta}.
\]

---

Fig. 1. An example of the data processing procedure taken from \( L_S = 30.6° \) in MY 25: brightness temperatures at the center of the CO\(_2\) 15-\(\mu\)m band sampled in the Southern Hemisphere along one orbit (dots); data same as the above, but binned in latitude increments of 0.5° with an offset of 10 K (filled circles); latitudinal segment spanning 32° ± 16° extracted from the data above, with an offset of 20 K (open circles); linear trend fitted to the latitudinal segment (solid line); and residual after subtracting the linear trend from the latitudinal segment, with an offset of 190 K (crosses). This residual is converted to a power spectrum.

Fig. 2. Numbers of the spectra that were calculated from (upper) single scan and (lower) double scan radiances and averaged in each latitudinal band and season.
Equations (1)–(3) yield

\[ \Phi_{PE} = \frac{g^2}{2N^2T^2} \Phi_{TT}, \]

by which \( \Phi_{TT} \) is converted to \( \Phi_{PE} \). The \( \bar{T} \) for each latitudinal region and season is calculated from TES-retrieved temperatures (Conrath et al. 2000) at the 0.50-hPa pressure level, where the contribution function peaks, distributed through PDS. The \( N \) is calculated from the combination of the temperatures at 0.50 and 0.64 hPa.

4. Results

Figure 3 shows the wavenumber spectra of potential energy \( \Phi_{PE} \) in 5 latitudinal bands as a function of season during MY 24–27. The wavelengths range from 64 to 957 km for latitudes 32°S, 0°, and 32°N, while they
range from 64 to 479 km for 64°S and 64°N. The spectral densities generally decrease with wavenumber and the slopes tend to be flatter at smaller scales. Some of the spectra steepen near the large-scale end with slopes from ~2 to ~3, while at smaller scales the slopes are usually flatter than ~1.

The contribution of measurement error seems to be negligible. The noise level $N_{TT}$ in the temperature spectrum in each latitudinal region and season is estimated by

$$N_{TT} = \frac{(\sigma_T / \sqrt{M})^2}{\Delta k},$$

(5)

where $\sigma_T$ is the RMS measurement error in brightness temperature, $M = 2$–5 is the mean number of observations within one 0.5° grid, and $\Delta k = 0.016$ km$^{-1}$ is the bandwidth corresponding to the Nyquist wavelength of 1° in latitude. The $\sigma_T$ is 0.4 K for the target temperature of 270 K and is 1.5 K for 150 K (Christensen et al. 2001); the $\sigma_T$ for each observed temperature is estimated by linearly interpolating the $\sigma_T$ values for 270 and 150 K. The $N_{TT}$ calculated by (5) is converted to the noise level in the potential energy spectrum, denoted by $N_{PE}$, in the same way as (4). The $N_{PE}$ is typically 10–30 m$^2$ s$^{-2}$ (km$^{-1}$)$^{-1}$ and generally much smaller than $\Phi_{PE}$. Exceptions are some of the spectra during MY 24, $L_S = 180°$–360°, in which $\Phi_{PE}$ is comparable to the $N_{PE}$ of ~10 m$^2$ s$^{-2}$ (km$^{-1}$)$^{-1}$ near small-scale ends. The contribution of measurement error is considered negligible also on the ground that most of the spectra have roughly constant downward slopes near small-scale ends and do not seem to flatten into the noise floor.

The spectra of the latitudinal segments that have been averaged in each season were also calculated (not shown): mesoscale structures have been averaged out but background structures remain in these segments. The calculated spectral densities are smaller than those shown in Fig. 3 by more than one order of magnitude. This ensures that the influence of background temperature structures on the spectra is negligible.

Figure 3 indicates that latitudinal tendency changes regularly with season. Around equinoxes ($L_S = 330°$–$0°$–30° and 150°–210°), the highest power is observed at 64°N and 64°S. Around solstices ($L_S = 60°$–120° and 240°–300°), the power peaks at 64°N or 64°S in the winter hemisphere, followed by 32°N or 32°S in the same hemisphere. On the other hand, seasonal variation in the power at each latitude tends to be obscured by large interannual variability; the power at small scales is suppressed in MY 24 and enhanced during the period from MY 25, $L_S = 210°$ to MY 26, $L_S = 120°$.

We also studied longitudinal dependence that may arise from the storm tracks around longitudes 150°–330°E (Hinson and Wilson 2002) or 200°–320°E (Banfield et al. 2004) in the southern mid- to high latitudes in winter. Figure 4 compares spectra averaged independently over longitudes 0°–180° and 180°–360°E in the Southern Hemisphere. The power is systematically larger at 180°–360°E than at 0°–180° during $L_S = 0°$–180°, being in harmony with the expectation. Such a longitudinal dependence is difficult to explain by artifacts and thus suggests the reliability of the analysis. No systematic tendency was found in the Northern Hemisphere (not shown), although Banfield et al. (2004) suggested that storm tracks will exist around 200°–320°E in the northern high latitudes.

5. Spectra of the terrestrial stratosphere

Here energy spectra of the terrestrial stratosphere are obtained with a similar method and compared to the Martian spectra. The data used are the atmospheric temperatures retrieved from nadir radiances taken by the Tropospheric Emission Spectrometer (TES) on NASA’s Earth Observing System (EOS) satellite Aura (hereafter referred to as the Aura TES, Beer et al. 2001; Beer 2006). The retrieved temperatures have vertical resolutions of 10–15 km that are similar to those of brightness temperatures, and thus the results would be roughly equivalent to those from nadir radiances, although the retrieval process improves the vertical resolution to some extent.

Aura was launched into a polar sun-synchronous orbit in 2004. The Aura TES is a Fourier transform spectrometer covering the spectral range of 650–3050 cm$^{-1}$. The products of the Aura TES are vertical concentration profiles of several minor gases, atmospheric temperature profiles, and surface temperature and emissivity. The standard observations of the Aura TES are global surveys, which consist of a sequence of observations including a view of cold space, a view of the internal blackbody, two nadir scans, and three limb scans. Since such an intermittent sampling does not fit our mesoscale investigation, we used data taken during one of the special observations, “Pacific Ocean,” in which 125 successive nadir scans (~0.4° latitude interval) were obtained repeatedly in the midlatitudes over the North Pacific Ocean during the period from 11 April to 19 May 2006. The retrieved temperature data (level 2) were obtained from the NASA Langley Research Center Atmospheric Sciences Data Center.

An analysis method similar to that for MGS TES data is applied to the retrieved temperatures at 100.0 hPa (~16-km altitude) and 31.6 hPa (~24 km) in the latitudinal band from 30° to 55°N along each orbit.
Temperatures in this height region are expected to be free from the influence of clouds. Each one-dimensional data series, which consists of 64 data points, is converted to a power spectrum by a digital Fourier transform with a Welch window function after removing the mean and a linear trend. The 87 spectra obtained are averaged for each altitude. The gravest Fourier component is excluded from the result since it is affected by the choice of the trend function and the window function. Finally each temperature spectrum is converted to a potential energy spectrum using the background $T$ and $N$ calculated for the same altitude, with the adjacent upper layer (90.9 hPa for 100.0 hPa, and 28.7 hPa for 31.6 hPa) also being used in calculating $N$.

The calculated spectra, which cover wavelengths from 95 to 1425 km, are shown in Fig. 5. The spectral densities of $\sim 300 \text{ m}^2 \text{s}^{-2} (\text{km}^{-1})^{-1}$ at small scales are similar to those calculated from stratospheric wind and

\begin{figure}[h]
\centering
\includegraphics[width=\textwidth]{fig4.png}
\caption{Same as Fig. 3, but for the spectra in the Southern Hemisphere averaged independently over longitudes $0^\circ–180^\circ$ (dotted) and $180^\circ–360^\circ$E (solid).}
\end{figure}
temperature data taken from aircraft (Lilly and Lester 1974). Given the specified temperature error of \(0.5\) K, the noise level in the spectra is as small as \(50\) m\(^2\) s\(^{-2}\) (km\(^{-1}\))^\(-1\). The spectral densities at wavelengths shorter than \(1000\) km are in the midst of the range over which the Martian spectra scatter.

The slopes are flatter at smaller scales: the slope near the large-scale end is about \(-2\) for 100.0 hPa and about \(-1\) for 31.6 hPa, and the rest of the spectrum shows a slope of \(-0.5\) for both altitudes. The relatively flat spectra at small scales might be explained by the contribution of gravity waves: the faster vertical propagation (less dissipation) of shorter-horizontal-wavelength gravity waves will make the spectrum flatter in the middle atmosphere, as suggested by Koshyk and Hamilton (2001) based on high-resolution GCM results. The rapid decrease in energy with height near the large-scale end might be due to Charney–Drazin filtering, which eliminates synoptic-scale eddy activity (Koshyk and Hamilton 2001).

The Martian spectra (Fig. 3) and the terrestrial stratospheric spectra (Fig. 5) obtained via similar methods show common features in the slopes that are relatively flat and tend to steepen near large-scale ends. This might imply that the physical processes contributing to the spectra in these atmospheres are similar, although the data used for the terrestrial stratosphere are limited in space and time.

![Fig. 5. Potential energy spectra of the terrestrial stratosphere at 100.0 hPa (filled circles) and 31.6 hPa (open circles) calculated from temperatures remotely measured by the Aura TES in the latitudinal band of 30°–55°N over the Pacific Ocean.](image)

6. Discussion

Based on the understanding of the spectra of the terrestrial atmosphere (sections 1 and 5), the steeper spectral region that sometimes appears near the large-scale end might be attributable to the forcing by synoptic-scale disturbances. Note, however, that the occurrence of an enstrophy cascade of quasigeostrophic turbulence is uncertain from the observations over a restricted wavenumber range. The latitudinal tendency that the power is maximized in the winter hemisphere implies forcing by baroclinic instability. It has been suggested that eddies arising from baroclinic instability will have deep structures extending to 50-km altitudes or more in the Martian atmosphere where the tropopause is absent (Barnes 1984); the altitude where the contribution function peaks (\(\sim 23\) km) is within the depth of such baroclinic eddies. The horizontal wavelengths studied are much shorter than those of baroclinic eddies observed or theoretically predicted, which have zonal wavenumbers 2–4 (Barnes 1980, 1981, 1984).

The flatter spectral region that dominates smaller scales will not be a part of the energy-cascading inertial range of two-dimensional turbulence, since its slope is much flatter than \(-5/3\). Brightness temperature spectra may be different from atmospheric temperature spectra because of the finite width of the contribution function (section 2); however, the latter cannot be steeper than the former, because smaller eddies will tend to have smaller vertical scales and will be smoothed out more effectively when sounded by the MGS TES.

A more plausible explanation for the flatter spectral region is the contribution of gravity waves. Given the horizontal wind speed of \(U = 40–120\) m s\(^{-1}\) (Banfield et al. 2003) and \(N = 0.008\) s\(^{-1}\) at the 0.5-hPa level, the vertical wavelengths of topographically generated gravity waves are \(2\pi UN^{-1} = 30–90\) km; based on the shape of the contribution function (section 2), waves with such vertical wavelengths will produce brightness temperature fluctuations with amplitudes of 0.25–0.78 times the waves’ amplitudes. In the presence of clouds such as the polar hood, which covers the winter high latitudes (e.g., Smith 2004), the contribution function will be more peaked and thus the brightness temperature will become more sensitive to atmospheric temperature fluctuations.

Wind and temperature data in the terrestrial atmosphere collected on airplanes show that variances are larger over mountains than over oceans or plains probably because of topographically generated gravity waves. Nastrom et al. (1987) and Jasperson et al. (1990) showed that the increase in the variance over mountainous regions is greatest at wavelengths of 4–80 km,
resulting in spectral slopes flatter than $-1$ for wavelengths longer than ~60 km. Lilly and Lester (1974) obtained spectra of the stratosphere (13–20-km altitudes) over the mountains of southern Colorado, showing that the gravity wave contribution peaks at wavelengths of 20–30 km and that the spectra are mostly flat at wavelengths of 20–200 km and follow a $-3$ power law at shorter wavelengths. Given such observations, a major contribution of gravity waves at mesoscales on Mars cannot be ruled out. The enhancement in regions of strong baroclinicity (Fig. 3) implies that gravity waves are generated by westerly jets flowing over topography, frontal activity associated with baroclinic instability, or the temporal meandering of westerly jets. The enhancement around the storm tracks in the southern mid- to high latitudes (Fig. 4) is also consistent with the generation of gravity waves in baroclinic zones.

Another possibility for the flatter spectral region is that horizontally nonuniform distributions of atmospheric dust and/or clouds create mesoscale structures in the brightness temperature along the MGS orbits. Fine structures in the distributions of low-level dust and clouds have been observed in visible images (e.g., Kahn 1984). Dust and clouds are also known to exist at high altitudes up to ~50 km (e.g., Jaquin et al. 1986), although their mesoscale structures are unknown. Dust column thickness is enhanced in the summer Southern Hemisphere ($L_S \sim 270^\circ$), while cloud thickness is enhanced near the equator around $L_S = 0^\circ$–180° and near the winter Pole (Smith 2004). Such seasonal and latitudinal tendencies do not explain the results; however, the mesoscale filamentation of dust and clouds as a result of synoptic-scale disturbances will be enhanced in the high latitudes in winter and equinoxes. More works are needed to evaluate this possibility.

The interannual variability in the power will be related to the year-to-year difference in the background atmospheric state, which is manifested by the intermittent occurrence of planet-encircling dust storms. An unprecedented large storm was initiated at $L_S = 185^\circ$ in MY 25 (year 2001) and the dust opacity remained noticeably higher than the normal value until $L_S = 340^\circ$ (Smith 2004). Even after the dust was settling out, the effect of the dust storm on atmospheric temperature lasted until the middle of MY 26. Smith (2004) showed that temperatures increased at high altitudes during the storm and that temperatures decreased near the surface with relatively small changes at 1–2 hPa during the poststorm period. Both of these temperature changes increase the static stability of the atmosphere at low altitudes, thereby allowing gravity waves to propagate more easily. Radiative effects of dust transported to high altitudes and distributed nonuniformly during the dust storm will also increase the power. Such changes in the background state will explain, at least partly, the enhancement of the power during the period from MY 25, $L_S = 210^\circ$ to MY 26, $L_S = 120^\circ$.

7. Summary

Wavenumber spectra of the atmospheric energy of Mars at mesoscales were investigated using the infrared radiance data taken by the TES onboard MGS. Latitudinal distributions of the brightness temperature at the center of the CO$_2$ 15-µm band along orbits were converted to power spectra, averaged for season, and converted to potential energy spectra. The obtained spectra tend to be flatter at smaller scales; the slopes are basically flatter than $-1$, but the spectra sometimes show steepening near the large-scale end with slopes from $-2$ to $-3$. The steeper spectral region at large scales might be attributable to the tail of the turbulent cascade from synoptic-scale disturbances driven by baroclinic instability, and the flatter region might be the contribution of gravity waves or the horizontally nonuniform distribution of dust or clouds.

In contrast to the “universal” spectrum of the terrestrial atmosphere (Nastrom and Gage 1985), the spectrum of the Martian atmosphere varies significantly with latitude and season and year by year. The latitudinal tendency changes regularly with season: around equinoxes the power has maxima at high latitudes of both hemispheres, while around solstices the power peaks in the high latitudes of the winter hemisphere followed by the midlatitudes of the same hemisphere. This suggests that eddies are generated preferentially in regions of strong baroclinicity. An enhancement of the power observed around the storm tracks in the southern mid- and high latitudes is consistent with eddy generation in baroclinic zones. Seasonal variation at each latitude band tends to be obscured by large interannual variability.

This study has made clear unique statistical characteristics of the mesoscale eddies in the Martian atmosphere. To fully understand the mechanisms of eddy energy injection and energy cascade, spectral analyses of synoptic- through planetary-scale eddies together with extended observations at mesoscales by the MGS TES or other instruments are needed. Comparison of the observed spectra with results from high-resolution Martian GCMs is also valuable for this purpose.

Acknowledgments. This study has been conducted using the MGS TES data and the Aura TES data distributed from NASA. The authors greatly appreciate the open data policy of these projects. The authors also thank two anonymous reviewers and Y. O. Takahashi for making valuable suggestions on this work.
REFERENCES


